Removal of a Constraint on the Composition of the Lunar Interior[1]

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Ringwood and Essene (1970) proposed that the CaO and Al_2O_3 contents of the deep interior of the moon must be less than 6% each. This constraint has been generally accepted and has led to models of the lunar interior that are dominated by ferromagnesium silicates. This constraint is invalid. High CaO and Al_2O_3 peridotites have broader intermediate-density (3.3–3.4 g/cm³) fields than the Ringwood-Essene 'lunar pyroxenite,' and the high-density phase occurs at higher pressure. Likewise, the gabbro-eclogite transformation pressure increases with Al_2O_3 content. The moon can have a thick plagioclase rich outer shell and a high Ca-Al interior.

The values for the moment of inertia and mean density of the moon place important constraints on the composition and mineralogy of the lunar interior. For example, basaltic achondrite meteorites or Apollo 11 basalts cannot be representative of the mean composition of the moon because they transform to dense garnet-bearing assemblages at shallow depth [*Wetherill*, 1968; *O'Hara et al.*, 1970; *Ringwood and Essene*, 1970]. *Ringwood and Essene* [1970] have generalized this situation and conclude that the overall contents of Al_2O_3 and CaO in the moon must be less than 6 and 5%, respectively, and that the moon, like the earth's mantle, is predominantly composed of iron-magnesium silicates. Strictly speaking, their petrologic arguments apply to the source region of the Apollo 11 basalts and not to the moon as a whole. Implicit in their conclusion, although not important for the present discussion, is the assumption that lunar temperatures are subsolidus throughout the moon. This limitation on the calcium and aluminum content has been adopted as a boundary condition by *Gast* [1972] and *Smith et al.* [1970] in their discussions of the composition of the lunar interior. *Wood et al.* [1970] and *Wakita and Schmitt* [1970] have proposed a chondritic interior. The purpose of this note is to point out that the now generally accepted constraint on the CaO and Al_2O_3 content of the lunar interior is not valid. A high CaO and Al_2O_3 interior can exhibit the desired property of surviving with acceptable densities to >30 kb or >700 km even at subsolidus temperatures. This removes the main motivation for considering the deep interior of the moon to be ferromagnesium silicates or chondritic in composition.

MacGregor [1970] has studied the effect of CaO on the stability fields of spinel and garnet peridotites. The situation is clearly summarized by him:

> The reaction pyroxene + spinel ⇄ garnet + olivine defines the boundary between low pressure spinel peridotite and high pressure garnet peridotite. Experimental determination in the four component system MgO-CaO-Al_2O_3-SiO_2 indicates that the CaO content is critical in defining the position of the reaction boundary. Where orthopyroxene is the only pyroxene present, increasing the CaO content decreases the pressure at which the reaction occurs; where both pyroxenes are present changing the CaO content has no effect; and where clinopyroxene is the only pyroxene present, increasing the CaO content increases the pressure of the reaction [*MacGregor*, 1970, p. 372].

The effect of CaO in this system is shown on a pressure-composition projection (Figure 1), which is modified from *MacGregor* [1970]. This is a projection of part of the join, pyrope-grossularite, and indicates the pressure at which

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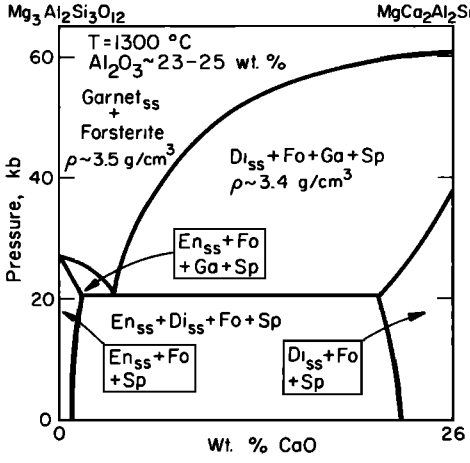

Fig. 1. Pressure-composition projection at 1300°C along the join pyrope-grossularite for a simplified peridotite [*MacGregor*, 1970].

the reactions occur as a function of CaO content. The Al_2O_3 content is relatively constant throughout the figure at 23–25 wt %. With increasing pressure, compositions involving two pyroxenes are the first to react to the high-pressure assemblage, and changes in the CaO content will not change the reaction pressure.

For very CaO poor, or CaO rich, compositions the transformation pressure depends on the CaO content. In the CaO poor field, increasing the CaO content decreases the pressure of reaction. In the CaO rich field, increasing the CaO content increases the pressure of reaction. The completion of the reaction to the dense assemblage initially decreases with CaO content, and then CaO > 3½ wt % (in this projection) increases. The transformation pressures for spinel to garnet pyroxenite or peridotite for a natural peridotite [*Ito and Kennedy*, 1967], 'pyrolite III' [*Green and Ringwood*, 1967*b*], and a 'model lunar pyroxenite' [*Ringwood and Essene*, 1970] agree well with this figure, even though their Al_2O_3 contents are only 2–5 wt %. The lunar moment of inertia data suggests that the density distribution is almost uniform. Clearly, this is more nearly the case for a CaO rich than for a CaO poor moon. A more general description of the phase relationships in the four-component system CaO-MgO-Al_2O_3-SiO_2 is given by *MacGregor* [1968].

Figure 2 shows the stability fields of the Ringwood-Essene 'model lunar pyroxenite' and, for comparison, the equilibrium boundary be-

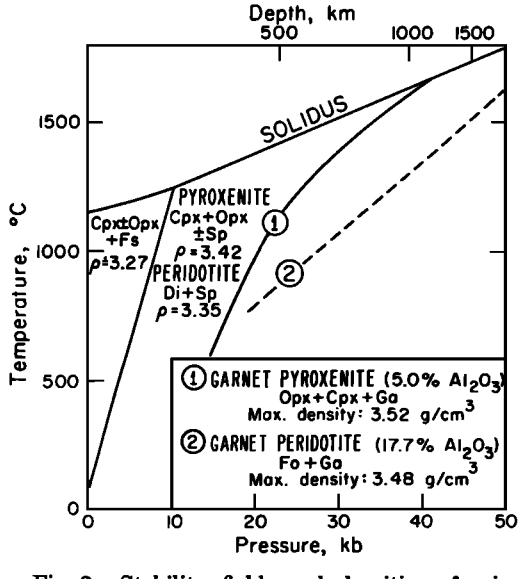

Fig. 2. Stability fields and densities of mineral assemblages displayed by Ca-Al deficient model lunar pyroxenite [*Ringwood and Essene*, 1970] and Ca-Al rich peridotite, 19.5% CaO, 17.7% Al_2O_3 [*MacGregor*, 1970].

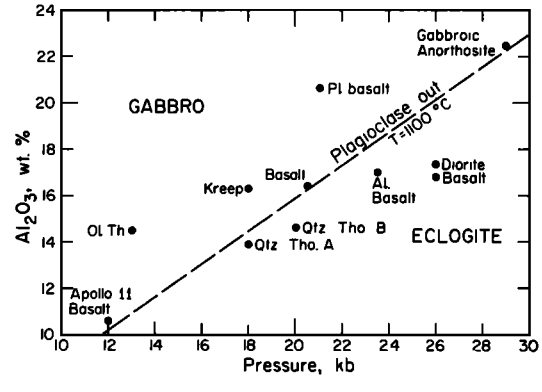

Fig. 3. The gabbro-eclogite reaction boundary as a function of Al_2O_3 content. Data were taken from the following sources: basalts, *Ito and Kennedy* [1970] and *Cohen et al.* [1967]; aluminous basalt, *Green* [1967]; quartz and olivine tholeiite, *Green and Ringwood* [1967*a*]; diorite and gabbroic anorthosite, *Green* [1970]; Apollo 11 basalt, *Ringwood and Essene* [1970]; KREEP and plagioclase basalt, *Ringwood and Green* [1972]. The materials on the high-pressure side of the dashed line mostly have more SiO_2 and less CaO than those on the low-pressure side. The CaO contents range from 7 to 13 wt %.

tween a spinel peridotite and garnet peridotite [*MacGregor*, 1970]. The latter contains 17.7-wt % Al_2O_3 and 19.5-wt % CaO. Note that the spinel peridotite field is much broader than the 'lunar' pyroxenite field, and the densities in both the spinel-bearing and garnet-bearing fields are lower than those in the comparable 'lunar' pyroxenite fields. The high-pressure field, for the CaO and Al_2O_3 rich assemblage, can be avoided entirely in the moon even at subsolidus temperatures.

The effect of Al_2O_3 content on the basalt → eclogite reaction also deserves some comment. Figure 3 shows the 'plagioclase out' boundary as a function of Al_2O_3 content at 1100°C. Although the Al_2O_3 content is not the only factor controlling the reaction pressure, there is a clear tendency for the basalt-eclogite boundary to increase in pressure as the Al_2O_3 content increases. The Na and Fe contents also affect the location of the boundary. Plagioclase, and therefore gabbro, is stable in a highly aluminous crust to great depth. A similar point has been made by *Ito and Kennedy* [1971]. *Gast* [1972] and *Anderson* [1973] have both suggested a Ca and Al rich outer shell of great [150–250 km] thickness.

Figure 4 gives the stability fields in three Ca and Al rich systems and, for comparison, a lunar temperature curve. Although none of these materials necessarily represent the composition of the deep interior, Figure 4 again makes the point that the CaO and Al_2O_3 contents of both the crust and the mantle cannot be required to be small. The intermediate-density field (ρ = 3.3–3.4 g/cm³) in these two high Ca-Al systems extends to greater depth than the 'model lunar pyroxenite' intermediate-density (ρ = 3.42 g/cm³) field.

The gabbroic anorthosite is similar in composition to that inferred to be appropriate by *Gast* [1972] and *Anderson* [1973] for the

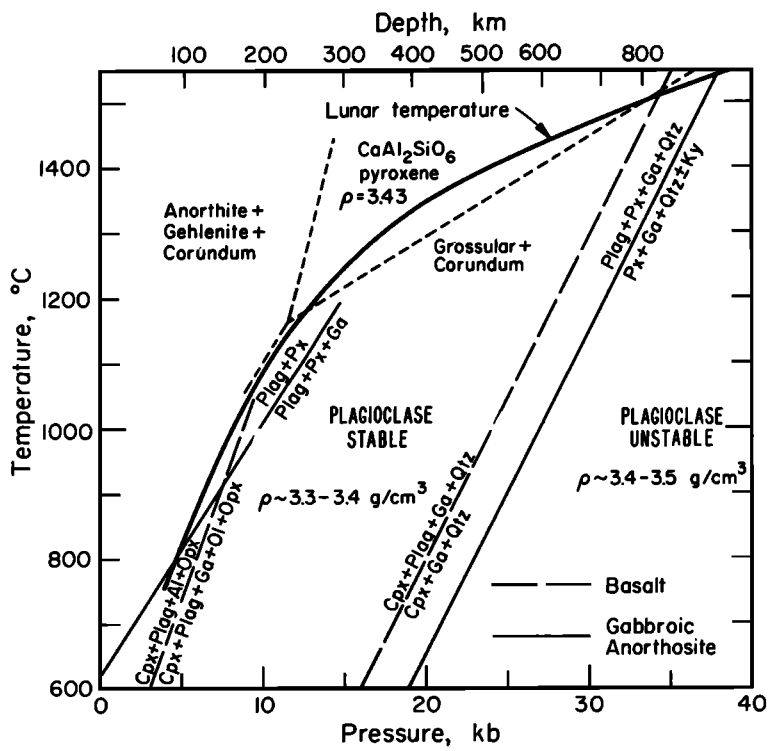

Fig. 4. Stability fields of three Ca-Al rich assemblages from *Ito and Kennedy* [1970] (basalt), *Green* [1970] (gabbroic anorthosite), and *Hayes* [1966] ($CaAl_2SiO_6$), in relation to a lunar temperature profile from *Hanks and Anderson* [1972]. *Ito and Kennedy* [1970, 1971] also suggest a very broad intermediate-density field in Ca-Al rich assemblage.

outer shell of the moon. For the lunar temperature profile shown, the low-pressure assemblage is stable between 80 and 370 km.

It has been proposed [*Anderson*, 1973] that the moon represents a high-temperature condensate from the solar nebula. In particular, it was proposed that the moon is enriched in the pre-iron condensates, which include Ca, Al, and Ti compounds and the refractory trace elements such as Ba, Sr, U, Th, and the REE. The residual crystals in this assemblage at lunar pressures are spinel, perovskite, and merwinite. The exterior shell, or crust, in this model is 250 km thick and is composed mainly of pyroxene and anorthite. It is enriched in Fe and the volatiles relative to the interior. Trace element distribution patterns suggest that the average composition of the crust is ~80% anorthosite and ~20% mare-type basalts. A thin garnet rich layer with high density and high seismic velocity can be stable in the crust (see Figure 4), depending on the temperature and composition.

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